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THE USE OF GENERAL CIRCULATION MODELS IN DETECTING CLIMATE CHANGE INDUCED BY GREENHOUSE GASES

by

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ABSTRACT

This article reviews some problems associated with the use of coupled atmosphere-ocean General Circulation Models (GCMs) in studies which attempt to detect a greenhouse-gas-induced signal in observed climate records. We show that model uncertainties affect both our predictions of how climate might change in response to greenhouse-gas (GHG) changes, and our estimates of the decadal- to century-time scale natural variability properties of the climate system. Knowledge of the latter are essential in order to make meaningful statements about when and even whether we could expect to detect a greenhouse-gas signal.

We show that GHG signal uncertainties are associated with errors in simulating the current climate in uncoupled and coupled climate models, the possible omission of relevant feedbacks, the non-uniqueness of the signal (due to the twin problems of the model's internally-generated natural variability and its sensitivity to initial conditions), uncertainties regarding the future GHG forcing and atmospheric GHG concentrations, and the so-called "cold start" error. Results from recent time-dependent greenhouse warming experiments are used to illustrate some of these points. We then discuss how energy-balance models, stochastically-forced ocean GCMs, and fully-coupled atmosphere-ocean GCMs have been used to derive estimates of decadal- to century-time scale natural variability, and consider some of the uncertainties associated with these estimates.

This review illustrates that it will be necessary to reduce both model signal and model natural variability uncertainties in order to detect a climate change signal and attribute this convincingly to changes in CO_2 and other greenhouse gases.

1. Introduction

In the last five years, a number of papers have been published in the scientific literature which have had as their focus the detection of greenhouse-gas-induced climate change in observed climate records. These studies can be divided into two general types—those dealing with both model and observed data, and those concerned with model data only. Investigations of the first type often have as their starting point the greenhouse-gas (GHG) signal predicted by a climate model, and then attempt to find this signal in observed records of surface temperature (Barnett, 1986; Barnett and Schlesinger, 1987; Santer et al., 1991, 1993a) or in records of temperature change in the lower stratosphere and troposphere (Karoly, 1987, 1989).

The results of such comparisons of model and observed data have been inconclusive (Wigley and Barnett, 1990). While they have failed to provide convincing statistical evidence for the existence of a GHG signal in the observations, they have also pointed out that there are many possible (and plausible!) explanations for such failure. Studies of this type have focused attention on problems of methodology (which statistical tools should we use in comparing model and observed data?), and on the difficulties involved in establishing a unique cause-and-effect link between changes in the climate of the last century and changes in greenhouse gases. They have also indicated that there are a number of areas in which the observed instrumental records of climate change and the model-predicted GHG signals show qualitative agreement (Houghton et al., 1990).

Studies dealing with model data only have had two purposes. The first purpose is to learn something about the *natural variability*[1] of the climate system. This provides us with information about the background "noise" of the climate system in the absence of any change in greenhouse gases caused by human activities. The second purpose is to identify climate variables which may be sensitive and highly specific indicators of GHG-induced climate change—in other words, variables which respond to changes in greenhouse gases in a unique way that cannot be confused with the

1. Natural variability can be defined as that portion of the total variability of climate which has nothing to do with changes in GHG concentrations, or with changes in other external factors which influence climate (e.g., the Sun's radiation output or volcanically-ejected dust), and is solely due to the internal dynamics of the atmosphere and ocean.

natural variability of climate, and that is also very different from the response to changes in other external factors.

The aim of this chapter is not to provide a comprehensive overview of previous detection studies and their principal findings. Instead, we will examine why detection of GHG-induced climate change is difficult, and address some of the problems associated with the use of model data in detection studies. The main issues which we address are:

- Model signal uncertainties
- Natural variability uncertainties
- The attribution problem

Before discussing these issues, we provide a brief introduction to climate models, and explain why models are essential tools in GHG detection studies. We also give some historical background to the different types of greenhouse warming experiment which have been performed.

2. Climate Models and Greenhouse Warming Experiments

There is no direct historical or paleoclimatic analog for the rapid change in atmospheric CO_2 which has taken place over the last century (Crowley, 1991). This means that we cannot use paleoclimatic data[2] or instrumental records in order to predict the regional and seasonal patterns and rate of climate change over the next century. We must therefore rely on numerical models of the Earth's climate system in order to make such predictions.

A large number of different numerical models have been used to study the effects of greenhouse gases on climate. The simplest of these consider the radiation budget at a single point on the Earth's surface. The most complex attempt to simulate the full three-dimensional circulation of the atmosphere and ocean. A typical fully-coupled ocean-atmosphere general circulation model (O/AGCM) generally divides the atmosphere and ocean into a number of discrete layers (extending from the bottom of the

2. By paleoclimatic data, we mean climatic data that can be inferred from such sources as tree rings, ice cores, lake varves, etc. (see, for example, Bradley and Jones, 1992), providing information about climate variability on time scales of centuries to thousands of years.

ocean to the top of the atmosphere), with each layer consisting of a two-dimensional grid of thousands of points. The model then solves equations for the transport of heat, momentum, moisture (in the atmosphere), and salinity (in the ocean) on this three-dimensional grid. A typical horizontal resolution in current O/AGCMs is 4° latitude x 5° longitude. Physical processes which occur on spatial scales smaller than the mesh of this grid (such as cloud formation) are *parameterized*—that is, their properties depend on the values of climate variables which are averaged over the 4° x 5° grid-cell. The bottom topography of the ocean and land surface orography are represented in a realistic way, but are smoothed to correspond to the resolution of the model. At a resolution of 4° x 5°, the Rocky Mountains generally do not exceed 2000 m, and orographic features which may be important for regional meteorological phenomena are not adequately resolved (see, e.g., Potter et al., 1993).

In O/AGCMs, as in the real world, the atmosphere and ocean communicate with each other, exchanging heat and momentum. The time scales of most atmospheric phenomena, such as frontal systems (with time scales of several days) and high pressure blocks (with time scales of weeks) are much faster than typical ocean time scales (of the order of centuries for the deep ocean circulation). The interaction between these fast and slow time scales can lead to a rich and complex spectrum of climate variability. It is essential to incorporate the coupling between the fast and slow components of the climate system in order to model natural variability and to understand how climate might change in response to gradually increasing concentrations of greenhouse gases (Hasselmann, 1988). Without an O/AGCM in which the ocean model is capable of realistically absorbing and redistributing heat from the atmosphere,[3] we will not have confidence in projections of the time evolution of the climatic response to GHG-forcing.

It is important to realize that the development of sophisticated O/AGCMs is a dynamic process. Such models evolve as computational speed and storage evolve, and as our understanding of the physics of the climate system improves. Until very recently, for example, most of our information concerning the possible climate response to GHG increases came from so-called *equilibrium response* experiments (see Schlesinger and

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3. This redistribution occurs both horizontally, via currents such as the Gulf Stream and Kuroshio, and vertically, in regions where water denser than underlying water masses sinks (e.g., in areas of the North Atlantic or the Antarctic Circumpolar Current) or less-dense water upwells.

Mitchell, 1987, and Mitchell et al., 1990). Such experiments generally used a relatively sophisticated atmospheric GCM, coupled to a much simpler model of the top layer of the ocean (the *mixed-layer*; usually the uppermost 50–100m). The experimental set-up involved instantaneously doubling the atmospheric CO_2 concentration (e.g., from 330 to 660 ppm), and then simulating the climate response over a period of 20–50 years. Because the mixed-layer of the ocean has a rapid response time (10–15 years), and since the longer time scales of the intermediate and deep ocean were neglected, such experiments allowed the climate system to reach a new equilibrium state within a relatively short time (10–20 years). The investigator then compared a sample of the model's new equilibrium climate with a sample from a control run without doubling of atmospheric CO_2 in order to learn something about the physics of the response.

In the real world, of course, CO_2 is increasing gradually and does not instantaneously double its atmospheric concentration. The more relevant question is how the climate system—including the deep ocean, with its longer time scales—will respond to slowly increasing GHG concentrations. It is only within the last few years that scientists have been able to address this question by performing *transient response* experiments with the O/AGCMs described above (e.g., Stouffer et al., 1989; Washington and Meehl, 1989; Cubasch et al., 1992). In a typical transient response experiment, an O/AGCM is forced by some scenario of how CO_2 and other greenhouse gases might change in the future. Scenarios which have been used in these experiments range from a simple linear increase in CO_2 (by 1% per year; Washington and Meehl, 1989) to the scenarios developed by the Intergovernmental Panel on Climate Change (IPCC; Houghton et al., 1990), which cover a range of optimistic and pessimistic assumptions about how world energy use and emissions might evolve over the next century.

Transient experiments add a new dimension to the detection problem. In addition to supplying information about the spatial pattern and seasonal features of the climate response, they also tell us something about the time evolution of the response on scales of decades to centuries. The time evolution is now of direct interest: the problem is to determine whether the trend describing the climate response (the *signal*) is significantly large relative to some distribution of decadal-to-century time scale trends which describes the natural variability properties (the *noise*) of the climate system.

3. Model Signal Uncertainties

As we have seen in the previous section, we are forced to rely on O/AGCMs in order to obtain information about the space-time properties of the climate response to GHG changes. The aim of this section is to consider the major uncertainties involved in these projections of GHG-induced climate change. We have partitioned these into six categories.

a. Errors in Simulating Current Climate in Uncoupled Models

Our confidence in the predictive capability of O/AGCMs when used in greenhouse warming experiments must be diminished by the knowledge that their individual atmospheric and oceanic components, when tested separately[4] to see how well they represent the current climate, still show systematic errors. A number of recent studies documenting the performance of atmospheric GCMs (Gates et al., 1990, 1992; Boer et al., 1992) have shown that, although model performance has generally improved over the last decade, all atmospheric models still have systematic errors in their simulation of the current climate (Gates, 1992). The ocean GCMs presently in use also have systematic errors (e.g. Maier-Reimer et al., 1993). Their validation is more problematic due to the difficulty of obtaining sufficient observed data, particularly for the intermediate and deep ocean.

b. Errors in Simulating Current Climate in Coupled Models

Even if an atmospheric GCM and an oceanic GCM, when tested separately, performed perfectly in simulating the present climate, there would be no guarantee that they would be equally successful when coupled together. In fact, experience shows that the interactive coupling of atmosphere and ocean GCMs generally leads to a phenomenon known as *climate drift*—that is, the tendency of the climate system to drift

4. By "testing separately", or "running in uncoupled mode", we mean that the atmospheric GCM is driven by the observed record of monthly-mean sea-surface temperature (SST) and sea-ice changes, for example over the period 1979–1988, rather than by the SST and sea-ice changes predicted by an ocean model (which will have their own sources of error). This enables us to isolate errors which are due to the atmospheric model only. A similar procedure is used to evaluate errors in the ocean model component.

into a new and unrealistic mean state (Gates et al., 1984; Washington and Meehl, 1989).[5]

We probably would not have much confidence in the predictive skill of the model if this new mean state were used as the starting point for a greenhouse warming experiment. In order to circumvent this problem, modelers usually use techniques known as *flux correction* or *flux adjustment*. This is a way of ensuring that the coupled model maintains a realistic mean state (Sausen et al., 1988). In a typical flux-corrected coupled model, the surface fluxes from the atmosphere into the ocean (e.g., of heat, net freshwater flux, and in some cases wind stress) and from the ocean into the atmosphere (e.g., of SST) are corrected, both spatially and over the seasonal cycle. Intuitively, one can think of these corrections as anomaly fields which are added to the computed fluxes, enabling the atmosphere and ocean to receive the fluxes that they need (rather than the uncorrected, erroneous fluxes that they get!) in order to maintain a stable climate.[6]

At the present state of development of coupled models, the flux corrections which must be made are sometimes as large as the flux changes predicted in greenhouse warming experiments, particularly in areas of strongly non-linear dynamics (sea-ice margins and areas of deep oceanic convection). Thus flux correction introduces an additional area of uncertainty in greenhouse warming experiments. Scientists are currently working to obtain a better understanding of the physics of atmosphere-ocean coupling, in order to reduce the magnitude of these corrections, and eventually to remove the need for an engineering solution to a scientific problem.

c. *Omission of Relevant Feedbacks*

Let us assume that we have an O/AGCM which realistically simulates the present climate without relying on any form of flux correction. Would this be a guarantee that the model would successfully predict the climate response to increasing GHG concentrations? The answer is, "probably not." Successful simulation of the present climate is probably a necessary, but not sufficient condition to ensure

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5. For example, ocean temperatures which are much colder than those observed on average, or an unrealistic distribution of sea-ice.

6. Note that once the flux corrections have been calculated, they remain invariant from year-to-year in a control run or experiment performed with the coupled model.

successful simulation of future climate. To be confident that our model has predictive skill on time scales of decades or longer, we would have to be sure that it incorporates all of the physics and feedback mechanisms that are likely to be important as GHG concentrations increase.

There are a number of reasons why it is difficult to feel confident that we have not forgotten anything important. We know, for example, that the feedbacks between clouds and the surface radiation budget are poorly understood. Cloud-radiation feedbacks involve such factors as the height, thickness, percentage coverage, and optical properties[7] of clouds. Recent studies have shown that different schemes for parameterizing cloud formation processes can lead to substantially different results in greenhouse warming experiments (Mitchell et al., 1989).

Numerous other examples are possible. Thus we know that O/AGCMs lack an interactive biosphere and treat surface hydrology in a relatively crude way. Most models do not explicitly consider the radiative effects of aerosols, or of greenhouse gases other than CO_2 and water vapor. They do not incorporate an interactive carbon cycle model, so it is difficult to determine whether a CO_2-induced change in climate could influence the uptake of atmospheric CO_2 by the deep ocean, and hence feedback on the climate change. In summary, therefore, we hope that current O/AGCMs incorporate all of the important physics and feedback mechanisms necessary to model the effect of increasing GHG concentrations on climate—but we cannot guarantee this.

d. Non-Uniqueness of Model GHG Signal

In any transient experiment with a fully-coupled O/AGCM, the model's own internally-generated natural variability will be superimposed on the true time-dependent GHG signal (Santer et al., 1993b). In the presence of substantial natural climate variability, the GHG signal is not uniquely defined. This is illustrated in Figure 1, which shows the time evolution of zonally-averaged annual mean surface air temperature changes in a 100-year greenhouse warming experiment recently performed with a coupled O/AGCM (Cubasch et al., 1992). In this experiment, the model was forced by time-varying GHG concentrations specified in the IPCC Scenario A ("Business-as-Usual"; Houghton et al., 1990).

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7. For example, the size distribution of water droplets or ice particles within the cloud, or the number of cloud condensation nuclei.

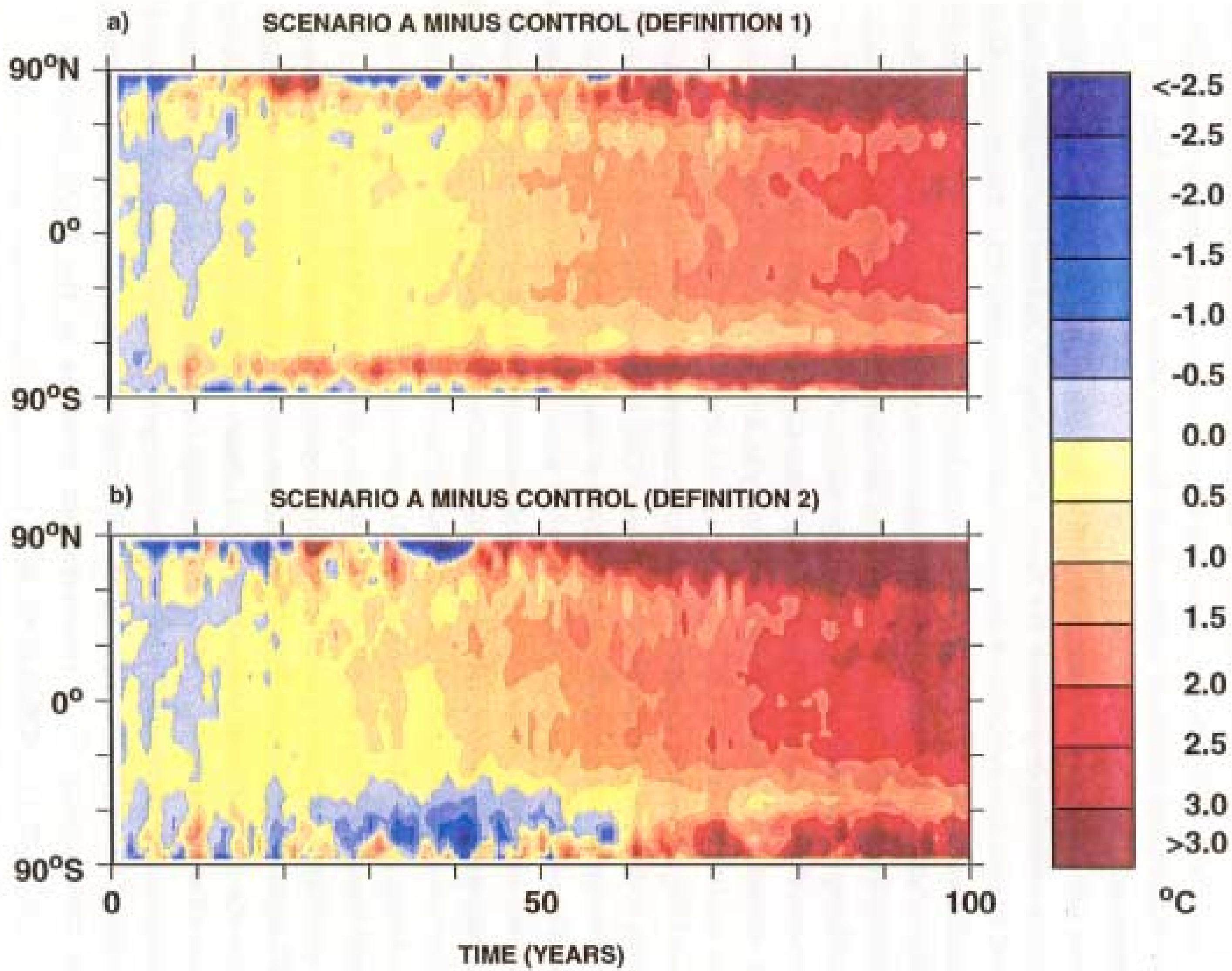

Fig. 1: Time evolution of changes in zonally-averaged annual mean 2m temperature in the 100-year Scenario A experiment performed by Cubasch et al. (1992). Changes are expressed relative to the average of the first 10 years of the control run (Definition 1; panel 1a) or the instantaneous state of the control run (Definition 2; panel 1b). The space-time evolution of the signal differs for the two definitions.

The two panels of Figure 1 show different definitions of the signal. In the upper panel, the changes in surface air temperature in the experiment have been defined by subtracting the average pattern of surface air temperature in the first 10 years of the control run. In the lower panel, the changes are defined relative to each individual year of the control run.[8] Although both definitions show the same qualitative picture of a slowly-emerging greenhouse warming signal, with the largest temperature changes at high latitudes in both hemispheres, the precise details differ. These differences are due to the model's natural variability in the control run, and/or the model's residual drift after flux correction (see Figure 2).

A further complication is the so-called *initial condition* problem. The experiment shown in Figure 1 commenced with a GHG concentration equivalent to that which existed in the atmosphere in 1985. Even in 1985 we had only limited observations about the climatic state[9] of the ocean, particularly the intermediate and deep ocean. This has the consequence that the experiment started from an ocean state which did not exactly correspond to the "true" ocean state which existed in 1985 but was imperfectly observed. Obviously, as one goes further back in time, sparser observations make it increasingly difficult to reconstruct a three-dimensional picture of the ocean's temperature and salinity.

It is only within the last few years that we have started to realize the implications of our imperfect knowledge of such initial conditions. While the pioneering work of Lorenz (1984) illustrated that the results of simple, "three equation" climate models are highly sensitive to initial conditions, such ideas have only recently been tested in the context of climate change experiments with O/AGCMs. Recently, Cubasch et al. (1993) performed a suite of three greenhouse warming experiments with the Hamburg O/AGCM. The three transient experiments forced the coupled model with identical increases in greenhouse gases (the equivalent CO_2 concentrations from 1985–2035 specified in the IPCC Scenario A), but each experiment started from different initial conditions. The initial conditions were three different "snapshots" of the Cubasch et al. (1992) 100-year control run (taken at years 30, 60 and 90). Together with the original 100-year Scenario A experiment performed by Cubasch et al. (1992), this

8. In other words, year 1 of the experiment minus year 1 of the control, etc.

9. For example, in terms of the three-dimensional structure of temperature and salinity.

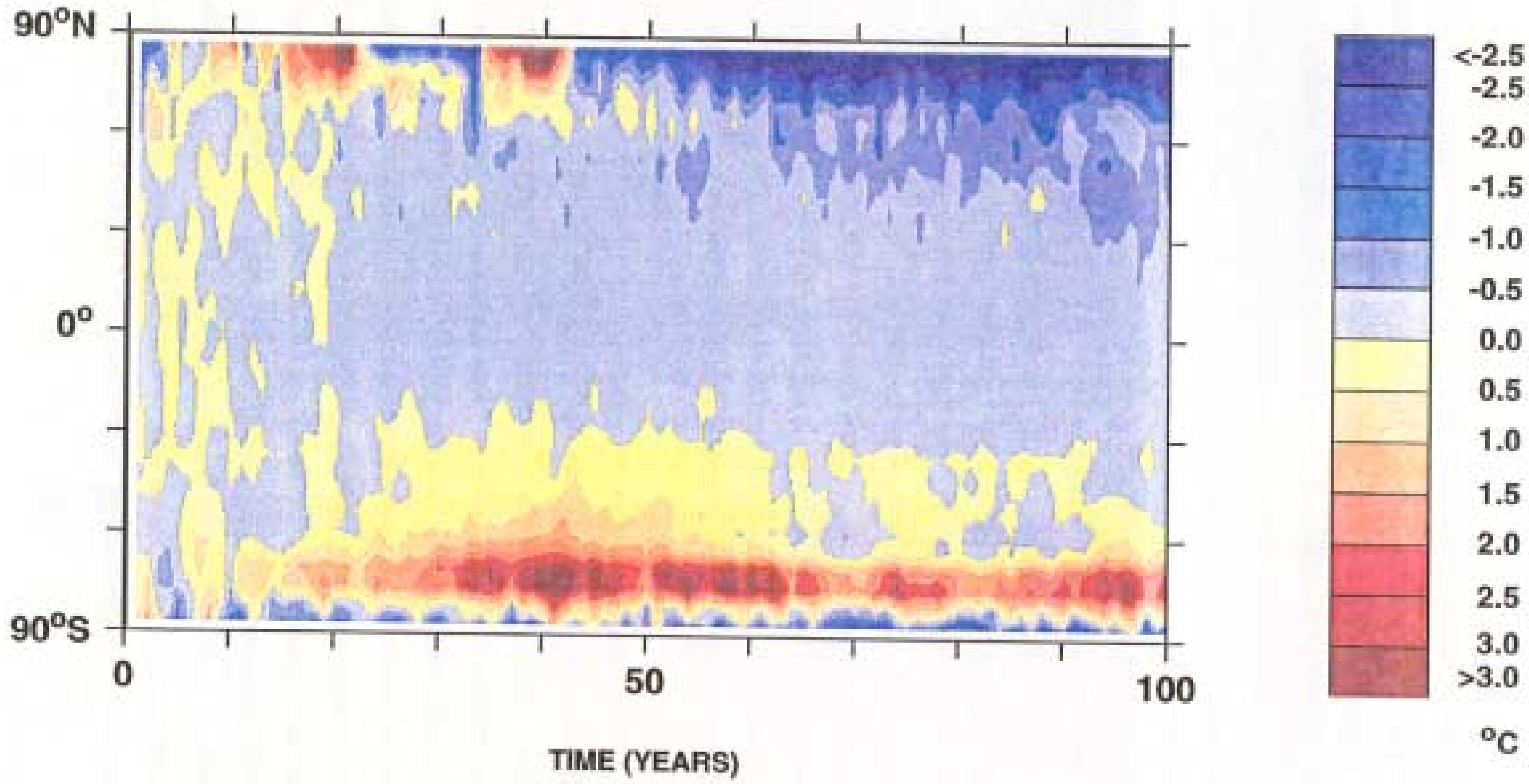

Fig. 2: Time evolution of changes in zonally-averaged annual mean 2m temperature in the 100-year control integration performed by Cubasch et al. (1992). Changes are expressed relative to the average of the first 10 years of the control run. Note the large variability at high latitudes in both hemispheres.

suite of four experiments provides some insights into the sensitivity of the greenhouse warming signal to the initial conditions of the climate system.

We found some notable differences in the space-time structure of the surface temperature signal in the four experiments. These are illustrated in Figure 3, which shows the results of an Empirical Orthogonal Function (EOF) analysis of surface air temperature in the 100-year Cubasch et al. (1992) control run. EOF analysis is a statistical tool which can be used to extract the dominant modes of variability from a large data set with observations at many points in space and time. We extracted the first two modes of variability from the original Cubasch et al. (1992) Scenario A experiment (the first two EOFs, which in this analysis are spatial patterns), and then projected the surface air temperature data from this experiment and the three initial condition experiments onto these two patterns. The results tells us something about how each of the four experiments evolves over space and time.

Each symbol on Figure 3 represents one year of the four greenhouse warming experiments. If the four experiments evolved randomly in space and time, Figure 3 would consist of a random distribution of points. Clearly, this is not the case. There is some *temporal coherence* to the results—in each experiment, the state of the surface temperature in year t bears some relation to the state of surface temperature in year $t - 1$, a reflection of the thermal inertia of the ocean.

If we assumed that the four greenhouse warming experiments were not sensitive to the initial conditions, we would expect them to evolve in a similar way over space and time. Their trajectories (plotted in EOF 1–EOF 2 space) would be very similar. However, Figure 3 shows that the surface temperature signal in the four experiments—each performed with the same model and the same GHG increases—can evolve very differently over space and time. This suggests that (even for a single coupled atmosphere-ocean GCM!) we might have to perform a large number of transient greenhouse warming experiments in order to obtain a good idea of the statistical properties of the climate system's response to GHG increases. The same argument obviously applies to the natural variability properties simulated in a control run.

e. Time Evolution of the Forcing

A further uncertainty is that we have no convenient crystal ball with which we can peer into the future and see how atmospheric GHG concentrations will change over the next 50–100 years. This uncertainty regarding the forcing has at least two

2M TEMPERATURE DATA: PROJECTIONS ON SCENARIO A EOFS 1 + 2

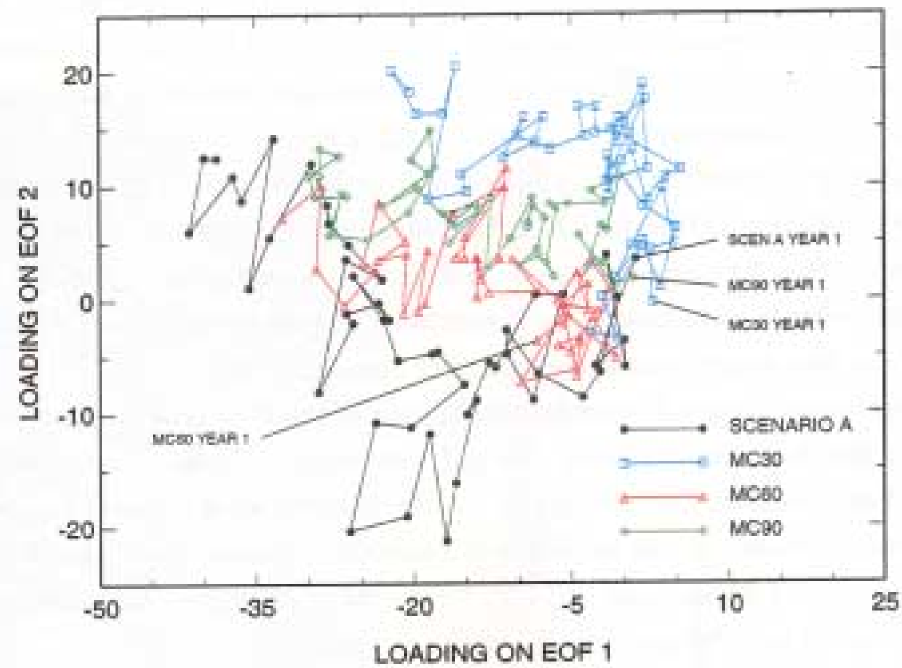

Fig. 3: Projection of the annually-averaged 2m temperature anomaly fields from four separate greenhouse warming experiments onto the Empirical Orthogonal Functions (EOFs) 1 and 2 of the Scenario A experiment performed by Cubasch et al. (1992). Each 50-year experiment used exactly the same greenhouse-gas forcing (the changes in equivalent CO_2 concentration from 1985–2034, as specified in the IPCC Scenario A), but started from different initial conditions of a 100-year control run. The initial conditions for the four experiments were taken from the beginning of the control run (for the Scenario A experiment), and then at years 30, 60, and 90 of the control run (for the MC30, MC60, and MC90 experiments, respectively). Each symbol represents one year of one experiment. For each experiment, the symbols are linked in order to show the evolution over time and in the 'space' of the first two EOFs. Although all four experiments have a general warming trend (which in this analysis shows up as a tendency for the warming signal to evolve towards the upper left-hand corner of the diagram), their trajectories in space and time are very different. This sensitivity to the initial conditions suggests that we might have to perform a large number of greenhouse warming experiments in order to obtain a good idea of the statistical properties of the climate system's response to greenhouse-gas forcing.

aspects—the difficulty of predicting future GHG emissions, and uncertainies regarding the global carbon cycle, which will determine how the emitted CO_2 is partitioned and cycled between the atmosphere, ocean, and biosphere.

A simple example shows how such forcing uncertainties are translated into uncertainties regarding the magnitude and space-time evolution of the signal. Recall that Figure 1a illustrated the zonally-averaged annual mean surface air temperature changes for the Cubasch et al. (1992) Scenario A experiment performed with the Hamburg O/AGCM. Figure 4 shows the corresponding temperature changes from an experiment with a lower level of GHG forcing (Scenario D). In Scenario D, some relatively optimistic assumptions are made regarding the reduction of GHG emissions after the year 2000 (see Houghton et al., 1990). A comparison of these two figures shows that there are differences both in absolute magnitude[10] and in the space-time evolution of the surface temperature signal in the two experiments.

f. *Cold Start Problem*

The final area of uncertainty in defining a transient GHG climate signal with an O/AGCM is the so-called *cold start* problem. This problem is related to the experimental set-up. Experiments such as those performed by Cubasch et al. (1992) were started with the atmosphere and ocean at equilibrium with respect to an equivalent CO_2 concentration of 360 ppm, which approximately corresponds to 1985 concentrations. In the real world, of course, there have been substantial changes in GHG concentrations *before* 1985. The neglect of this previous history of the GHG forcing means that we are neglecting any GHG-induced warming of the ocean which has taken place prior to 1985. This is the "cold start" error—the ocean has not been "warmed up" before the start of the experiment. Using simple linear models, Hasselmann et al. (1992) estimated that this error may be as large as 0.4°C after 50 years of the Cubasch et al. (1992) Scenario A experiment. A further analysis by Fichefet and Tricot (1992), using an upwelling-diffusion energy-balance model, yielded a cold start error amounting to roughly 15% of the warming signal if the integration starts in 1990 instead of in 1765.

Obviously, it would be more realistic to start a transient greenhouse warming experiment with atmospheric GHG concentrations appropriate to 1900 or even earlier.

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10. The globally-averaged annual temperature change after 100 years is over four times larger in Scenario A than in Scenario D (2.6°C versus 0.6°C, respectively).

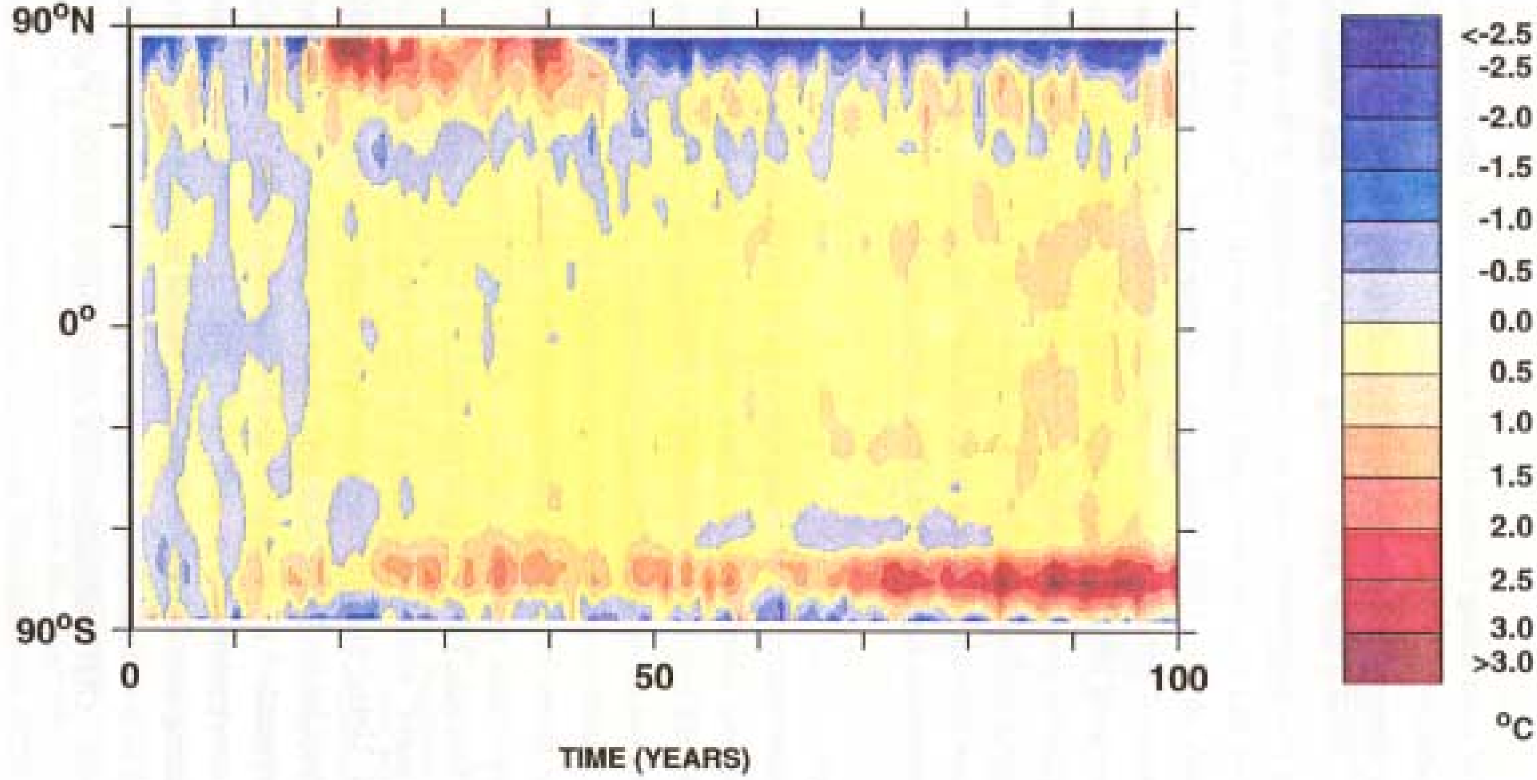

Fig. 4: Time evolution of changes in zonally-averaged annual mean 2m temperature in the 100-year Scenario D experiment performed by Cubasch et al. (1992). Changes are expressed relative to the average of the first 10 years of the control run. Note the differences in absolute magnitude and space-time evolution of the signal relative to Scenario A (Fig. 1a).

A rigorous investigation of the cold start error will require such an experiment. At present, however, this is impractical since even a single 200-year O/AGCM integration on a state-of-the-art supercomputer consumes many months of computer time. It is also worth noting that many such experiments would be required. Due to the twin problems of model-generated natural variability and sensitivity to initial conditions (see Section 3d), a single experiment cannot provide a single, definitive estimate of the magnitude of the cold start error.

4. Natural Variability Uncertainties

The previous section discussed the uncertainities associated with the models which were used to define the climate change signal likely to result from future changes in atmospheric GHG concentrations. But the climate change signal is only one part of the detection problem. In order to say something meaningful about when (or even whether!) we could expect to detect a GHG signal, we need to have good estimates of the natural variability of the climate system on time scales of decades to centuries. Information on natural climate variability can be derived from three sources:

a. Instrumental Records

In an ideal world, we would have observed surface air temperature over the last 1,000 years, using an observing strategy and a network of observing stations well-suited to the detection of a GHG signal and the separation of this signal from century-time scale natural fluctuations (e.g., the Little Ice Age). Our "ideal strategy" would have specified that there should be no changes in the instrumentation used to record temperature,[11] that observing times, frequencies, or practices should not change, that the location and elevation of observing stations should not change, and that stations should not be located in or near rapidly-growing urban areas. Our "ideal network" would have had enough stations to obtain a reasonable record of temperature variations over the entire Earth's surface (at least on spatial scales of several hundred kilometers), with little or no change in the number of observing stations as a function of time.

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11. Or that there should be sufficient overlap between an old instrument and its replacement in order to calibrate the new instrument, and prevent discontinuities in the record.

Unfortunately, neither the ideal observing strategy nor the ideal observing network exist! All of the problems alluded to above make it difficult to reconstruct a homogeneous, spatially-complete picture of surface temperature changes over the last century (see, e.g., Folland et al., 1990; Jones et al., 1991). We can attempt to correct some of these errors, such as the effects of urban warming on temperature records (Jones et al., 1989, 1990). Other problems, such as the deterioration in the spatial coverage of the observing network as one goes further back in time, are essentially insoluble.

Recently, it has been suggested that satellite records of near-surface temperature may provide the solution to some or all of these problems (Spencer and Christy, 1990). While satellite-derived data are spatially complete, they measure temperature in the lower troposphere and not at the Earth's surface. A more serious failing is the short length of available record (a decade or less). Satellite data cannot provide us with information about the natural variability of the climate system on time scales of decades.

b. Paleoclimate Records

Changes in climate affect a wide range of biological, chemical, and geological processes. As a result, climatic information is naturally recorded in tree rings, ice cores, coral reefs, laminated sediments, etc. (e.g., Crowley and North, 1991; Bradley and Jones, 1992; Briffa et al., 1992). If we can understand the recording mechanism—for example, the process by which climate imprints itself on tree growth and annual ring formation—then we have the potential to unlock a wealth of climate information stored in paleoclimate records.

Unfortunately, unraveling the history of climatic variability contained in such records is not a simple task. For example, many types of tree are more sensitive to moisture stress than to temperature, or may respond to non-climatic factors (e.g., changes in management practices). This makes it difficult to extract a temperature signal from the noise introduced by the variations in other factors which affect tree ring width. More importantly, spatial coverage is poor for paleoclimate data which can resolve annual temperature variability, and it is difficult to date and cross-check the climate information extracted from different locations (e.g., land and ocean) or from different proxy sources. For these reasons and many others, scientists have been unable to use paleoclimate data in order to reconstruct a satisfactory, spatially-complete picture of climate variability over the past 1,000 years.

c. *Numerical Models*

Numerical models provide another means of investigating the decadal-to-century-time scale variability of the climate system. Some of the first model-based studies of natural variability used simple energy balance models (EBMs; e.g., Hasselmann, 1976). Such models generally solve equations for the heat balance of a highly-idealized representation of the Earth's atmosphere and ocean (Lemke, 1977). By forcing an EBM with white noise—for example, by heat flux anomalies which are essentially random in time,[12] and thus can be thought of as characteristic of day-to-day weather noise—it was possible to investigate the relationship between daily weather noise and the model's internally-generated variability on time scales of years to centuries. These early studies, together with more recent EBM studies by other groups (e.g., Wigley and Raper, 1990, 1991; Kim and North, 1991), have demonstrated that even simple EBMs can generate decadal-to-century-time scale surface temperature fluctuations as an integrated response to daily-time scale random weather fluctuations.

While EBMs successfully reproduce many details of observed surface temperature variability on the annual-to-decadal-time scales (Kim and North, 1991), they are not as useful on longer time scales, since they cannot explicitly simulate the horizontal and vertical transport of heat, salt and momentum necessary for an accurate representation of the ocean circulation. It is therefore necessary to use more sophisticated models in order to obtain information on century-time scale natural variability.

Ideally, it would be desirable to study the century-time scale natural variability of the climate system by performing an ensemble of long (more than 1,000 years) control runs with a fully-coupled O/AGCM. Due to computational restrictions, however, state-of-the-art O/AGCMs have generally been integrated for 100 years only (Stouffer et al., 1989; Cubasch et al., 1992). This is too short to obtain reliable information about the statistical properties of climate on the century time scale. Even if computational restrictions did not exist, we have the additional problem that at least some of

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12. But which have an amplitude, (and possibly also a spatial structure) typical of observed heat fluxes.

the variability exhibited by a fully-coupled O/AGCM may be spurious climate drift attributable to inadequacies in the flux correction scheme (see Section 3b). Distinguishing between residual drift and real natural variability of the coupled system is a difficult task (see Santer et al., 1993b).

This does not mean that we have to wait several years for the next generation of supercomputers before performing experiments which supply useful information about century-time scale natural variability. One possible answer is to extend the philosophy of noise-forced EBMs to its logical conclusion, and to force an uncoupled OGCM by white noise. This is computationally efficient (since the main computational burden in O/AGCM experiments is the atmosphere), which means that it is relatively inexpensive to integrate an uncoupled OGCM for several thousand years. The assumption underlying this type of experiment is that the ocean (with its very long time scales) is the most important player in determining the climate system's century-time scale natural variability, and that the atmosphere is a more or less passive "slave" whose behavior is relatively unimportant in terms of long time scale climate variability. The motivation is to see whether the ocean has preferred patterns and time scales of variability when forced by atmospheric weather noise.

This approach has recently been tested by Mikolajewicz and Maier-Reimer (1990) in a 3,800-year experiment. They forced the Hamburg OGCM with fresh water flux anomalies[13] which were white in time but had an amplitude and spatial structure characteristic of observed fresh water flux anomalies. The response of the ocean was extremely complex. Using advanced statistical techniques, it was possible to isolate different ocean modes of natural variability, each with its own characteristic time scale and spatial pattern. The dominant mode had a time scale of roughly 320 years and described the movement of large-scale salinity anomalies through the Atlantic via the model's *conveyor belt* circulation.[14] This mode shows up clearly in spectra of ice volume, mass transport, heat fluxes, and many other ocean variables (see Figure 5). It is interesting to note that temperature reconstructions from Greenland ice cores also have a dominant mode of variability with a time scale just in excess of 300 years

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13. Note that the fresh water fluxes from the atmosphere into the ocean are determined by the net balance between the processes of precipitation, evaporation, and surface runoff.

14. The term "conveyor belt circulation" refers to the horizontal and vertical movement of large water masses within and between ocean basins.

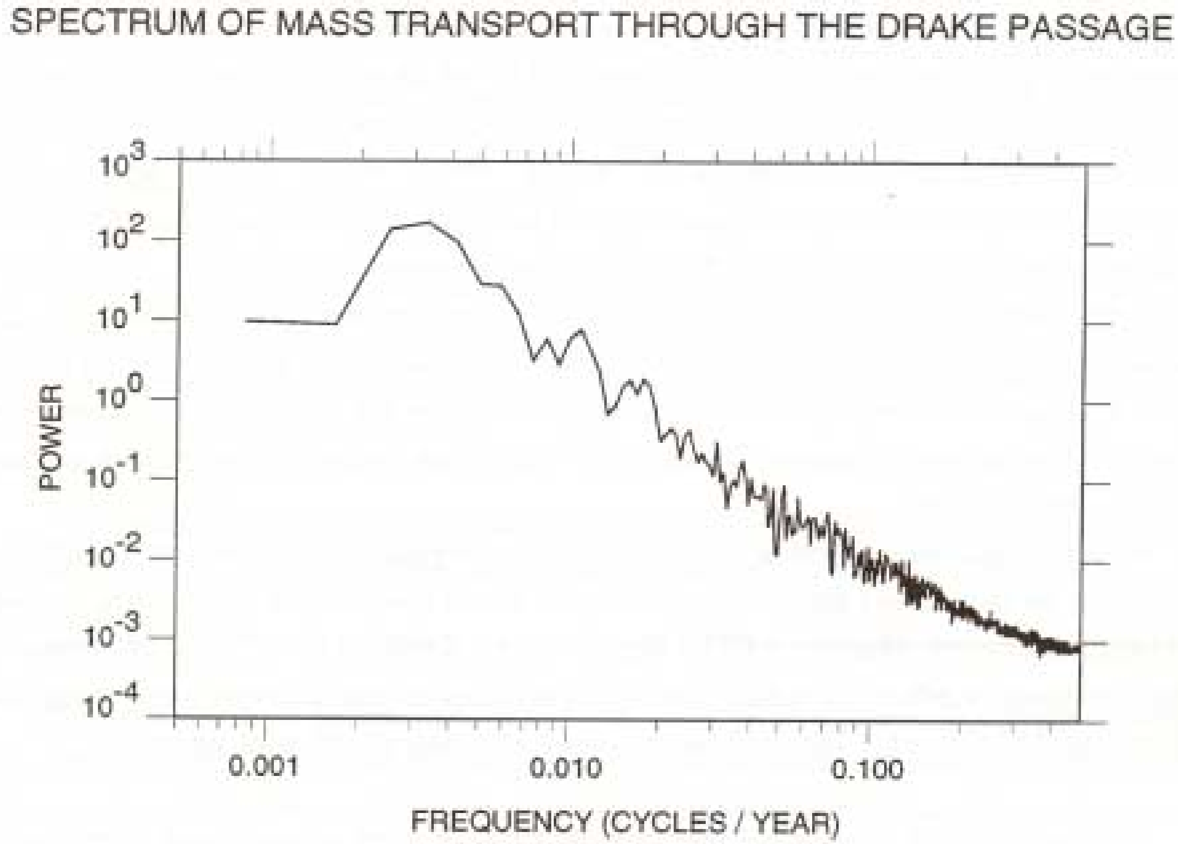

Fig. 5: Spectrum of mass transport through the Drake Passage from the 3,800-year 'ocean only' experiment performed by Mikolajewicz and Maier-Reimer (1990). Results are for a chunk length of 1,200 years, and are averaged over three non-overlapping chunks. Input time series were normalized. Note that the spectrum has maximum power at a period of approximately 320 years. The mode of variability associated with this spectral peak describes the movement of large-scale salinity anomalies through the Atlantic.

(Mikolajewicz and Maier-Reimer, 1991)—a tantalizing correspondence between the model and the real world, which deserves further investigation. Century-time scale modes of variability similar to that found by Mikolajewicz and Maier-Reimer have also been found recently by Mysak et al. (1993) in a stochastically-forced 2-D ocean circulation model.

One drawback with using noise-forced OGCMs for studying century-time scale natural variability is that we do not know whether they are truly representative of the long time scale variability likely to occur in a fully-coupled O/AGCM. Although the important dynamics of the ocean is reproduced, and the effective forcing by the atmosphere is probably adequately represented by white noise, the model does not include any feedback with the atmosphere. In the Mikolajewicz and Maier-Reimer experiment (1990), for example, wind stress, fresh water fluxes and surface temperature were fixed at their climatological monthly mean values,[15] so that the atmosphere was unable to respond to any change in oceanic circulation, and in turn modify the pattern or time scale of the dominant ocean variability modes. The fact that surface temperature was held fixed at its climatological values in this experiment also has the consequence that the information we are probably most interested in for GHG detection purposes—the century-time scale variability of surface temperature—is unavailable.

Without reliable estimates of decadal-to-century-time scale natural variability, we will not be able to say anything meaningful about how long it will take to detect a GHG signal, or even whether a GHG signal can be detected at all! It is important to make a concerted effort to reduce the uncertainties in our knowledge of long time scale natural variability. This can be done in a number of different ways:

• By attempting to validate the variability data from noise-forced "ocean only" experiments and fully-coupled O/AGCM control runs. Rigorous validation will involve making comparisons with appropriate paleoclimate data (Crowley and Kim, 1992). The development of a paleoclimate data set suitable for validation purposes will require a major international effort to date and cross-check the information from different geographical locations and different proxy sources.

• By exploring the sensitivity of results from noise-forced OGCMs to the horizontal and vertical resolution of the model and to the precise details of the forcing—e.g., the amplitude, correlation time and correlation scale of the forcing, whether the forcing is applied to the fresh water fluxes, wind stress, etc. (Mikolajewicz and Maier-Reimer, 1991; Barnett et al., 1993; Mysak et al., 1993).

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15. The spatially-coherent, temporally-white fresh water flux anomalies used to force the model were superimposed on the climatological mean fresh water fluxes.

• By performing many 100–200-year control runs with a single O/AGCM, with each experiment starting from different (but plausible) initial conditions of the climate system (see Section 3d), or a much smaller number of long integrations (more than 1,000 years). This will provide some insight into the statistical properties of the climate system on the century time scale.

• By trying to determine how much of the variability we see in control runs with fully-coupled O/AGCMs is *bona fide* natural variability of the coupled system, and how much is residual climate drift due to inadequacies in the flux correction scheme and in the physics of the coupling.

• By studying the model-dependence of natural variability results—for example, whether the modes of variability simulated by two totally different OGCMs are at all comparable.

The validation of model variability will be a difficult task. While the century-time scale variability in paleoclimate records reflects the response of the climate system to a complex mixture of external forcing factors (solar variability, volcanic and sulfate aerosols, etc.) and internally-generated variability, the variability simulated by an O/AGCM represents *only* the natural variability of the coupled ocean-atmosphere system. In order to validate model variability in a more meaningful way, it may be necessary to perform experiments in which a coupled model is forced by the past changes in volcanic aerosols and solar luminosity.

We will always have to live with model-based uncertainties in defining the regional and seasonal details of a GHG signal. The fidelity with which models simulate decadal-to-century time scale natural variability should be testable, however, if we can be clever enough to extract information from the silent biological, chemical, and geological witnesses to climatic change.

5. The Attribution Problem

Let us assume that we have actually managed to detect the GHG signal predicted by a model in the observed record of surface temperature changes. This means that we have used some statistical technique to compare the model signal with the observed data. The correspondence between the two is so striking that we conclude (on the basis of some statistical test at a prescribed level of significance) that our result could not be due to chance alone.

This result does *not* mean that we have established a clear causal link between changes in GHG concentrations and changes in surface temperature. In order to attribute the change in climate to the change in greenhouse gases, we would have to rule out all other possible explanations for the climatic change. We would have to demonstrate in a convincing way that changes in solar luminosity, volcanic aerosols, sulfate aerosols, or other external forcing factors could not have resulted in the observed surface temperature changes. We would also have to demonstrate that the internally-generated variability of the climate system on time scales of decades to centuries could not be confused with a slowly-evolving GHG signal. Finally, we would have to show that no combination of these external forcing changes or internal natural variability could explain the observed changes.

Given the uncertainties in our understanding of natural variability (see Section 4) and in our knowledge of the history of solar and volcanic forcing (and other forcing mechanisms), it is easy to see why attribution is a much more difficult task than detection. For example, evidence from experiments investigating the model response to changes in the solar constant suggests that the pattern of surface temperature change may be similar to the response pattern obtained in greenhouse warming experiments (Wigley and Jones, 1981). This result may be due to the fact that the surface temperature changes are at least partly due to feedback mechanisms (such as ice-albedo feedback) which respond to *both* GHG and solar forcing.

Another example is the vertical profile of temperature change (stratospheric cooling and tropospheric warming), a common feature of greenhouse warming experiments. Recent work by Santer et al. (1993b) with the Hamburg O/AGCM suggests that (at least in this particular model) the simulated natural variability pattern can look similar to the profile of vertical temperature change which the model predicts in greenhouse warming experiments. This would mean that natural variability could mimic a GHG signal.

Clearly, it will be difficult to solve the attribution problem if we use a single variable only, such as temperature changes at the Earth's surface. By considering a number of climate variables simultaneously, we will probably have a better chance of defining a climatic *fingerprint* which is unique to changes in greenhouse gases (Madden and Ramanathan, 1980; MacCracken and Moses, 1982; Hasselmann, 1993). The key to any fingerprint strategy is that we cannot use a hundred different climate variables simultaneously. We should focus on those variables for which suitable observed data exist, and which have high *signal-to-noise* ratios in the model data (in other words, variables which provide much more information about a GHG signal than about the model's own natural variability).

6. Conclusions

State-of-the-art O/AGCMs are the product of large research teams and represent an enormous investment in terms of time, money, and scientific effort. Running these models for long greenhouse warming experiments may require months of CPU time, even on the fastest supercomputers. In view of this investment and the complexity of the models themselves, there is a tendency to regard the results of greenhouse warming experiments conducted with O/AGCMs as being "engraved in stone". The reasons that we have given above suggest that this would be a mistake. State-of-the-art coupled models can give us internally-consistent pictures of a possible future climate, and can teach us about the physical mechanisms which are likely to be important in changing climate. We should regard them as instruments for making intelligent guesses about future climate, rather than as instruments for making definitive predictions.

We have seen that model uncertainties affect both our predictions of GHG-induced climate change and our estimates of decadal-to-century time scale natural variability. In order to detect a climate change signal and attribute this convincingly to changes in CO_2 and other greenhouse gases, it will be necessary to make progress reducing both signal and noise uncertainties.

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